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Late Pliocene upwelling in the Southern Benguela region

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Abstract

The Late Pliocene has been proposed as a possible analogue for understanding future climate change and for testing climate models. Previous work has shown that during the Pliocene the major upwelling systems were relatively warm, and that this meant they were either inactive, contracted, or were upwelling warmer waters than present. Here, we examine evidence from a site located on the margins of the modern Benguela Upwelling system to test whether the upwelling cells had migrated or contracted relative to present during the Pliocene.

We applied several organic geochemistry proxies and foraminiferal analyses to reconstruct the Pliocene history of ODP Site 1087 (31°28'S, 15°19'E, 1374m water depth), including the $U^{K_{37}}$ and TEX_{86} indices (for reconstructing sea surface temperatures), phytoplankton biomarkers concentrations and stable isotope ratios (for estimating export primary productivity, and for oxygen isotope stratigraphy), and planktonic foraminifera assemblage abundances (for inferring water mass changes). These proxies show that, between 3.5 and 3.0 Ma, the southern Benguela region was cooler than the northern Benguela region by 5°C, the latter being where the main upwelling cells are found today. From the multiproxy data obtained, we also infer that more extensive upwelling was present in the southern Benguela region during the Pliocene than at present, and that the Benguela Upwelling cells shifted northwards after the Pliocene epoch as a result of changes in the local wind field. We also find evidence that the Benguela Upwelling was sensitive to the pronounced cooling during the M2 and KM2 glacial stages, potentially associated with the expansion of sea ice and cooling in Antarctica in the Late Pliocene.

Keywords: Pliocene, Benguela Upwelling, Agulhas Leakage, M2, Antarctica, Southern Hemisphere, alkenones, GDGTs, foraminifera

1. Introduction

The Late Pliocene warm period, or the late Piacenzian, occurred between 3.3-3.0 Ma (Dowsett et al., 2012). This period has been studied extensively as a possible climate analogue for future warming (Dowsett et al., 1996), as climate conditions might have been similar to those predicted for the end of this century (IPCC, 2014). For instance, atmospheric CO₂ concentrations during the Late Pliocene have been reconstructed to be up to 450 ppmv, and thus lie close to those of end of 21st century climate predictions (Dowsett et al., 1996, 2012; Dowsett and Robinson, 2009; Henderiks and Pagani, 2007; Martínez-Botí et al., 2015; Seki et al., 2010). Within the overall warmth of the Pliocene, a pronounced excursion in the benthic $\delta^{18}\text{O}$ record marks the “M2 glaciation” at 3.3 Ma, which is seen as an early major cooling in global climate before the later onset of Northern Hemisphere glaciation from 2.6 Ma (De Schepper et al., 2009; Prell, 1984). To understand the climate system response to a warmer, higher CO₂ world (and the cooling events within this warm state), it is important to examine the evidence for regional and local responses in circulation change and biological productivity.

It is thought that during the Pliocene the major coastal and equatorial upwelling systems were either diminished in intensity, not present, or had a radically different mode of operation although more recent model reconstructions suggest an intensification of upwelling with future warmer temperatures (Brierley and Fedorov, 2010; Dekens et al., 2007; Etourneau et al., 2010, 2009; Fedorov et al., 2010; Harvey, 2000; Leduc et al., 2014; Rosell-Melé et al., 2014; Wang et al., 2015). The Benguela Upwelling system is one of the major upwelling cells in the modern global ocean (Boebel et al., 2003). It is located in the southeast Atlantic (Figure 1), and is divided into two regions: perennial upwelling in the northern and central Benguela, and seasonal upwelling in the southern Benguela (Boebel *et al.*, 2003). Previous work on the Pliocene-Pleistocene history of the Benguela Upwelling has focused on the northern and central Benguela region (Etourneau et al., 2010, 2009; Leduc et al., 2014; Marlow et al., 2001; Rosell-Melé et al., 2014), but less is known about the southern region. In the modern ocean,

waters offshore of the southern Benguela are also influenced by the intensity of the Agulhas Leakage, which brings warm and salty Indian Ocean waters into the Atlantic, playing a key role in heat and salt transport through the global ocean system (Beal et al., 2011; Gordon et al., 1992; Weijer et al., 2001). Model reconstructions suggest that the Agulhas Leakage could have been more vigorous during past warmer climates, and located in a similar position as today (McKay et al., 2012). However, other studies suggest a diminished Agulhas Leakage during the Pliocene because of reduced Indonesian Throughflow, which ultimately feeds the Agulhas Current (Karas et al., 2011a, 2011b).

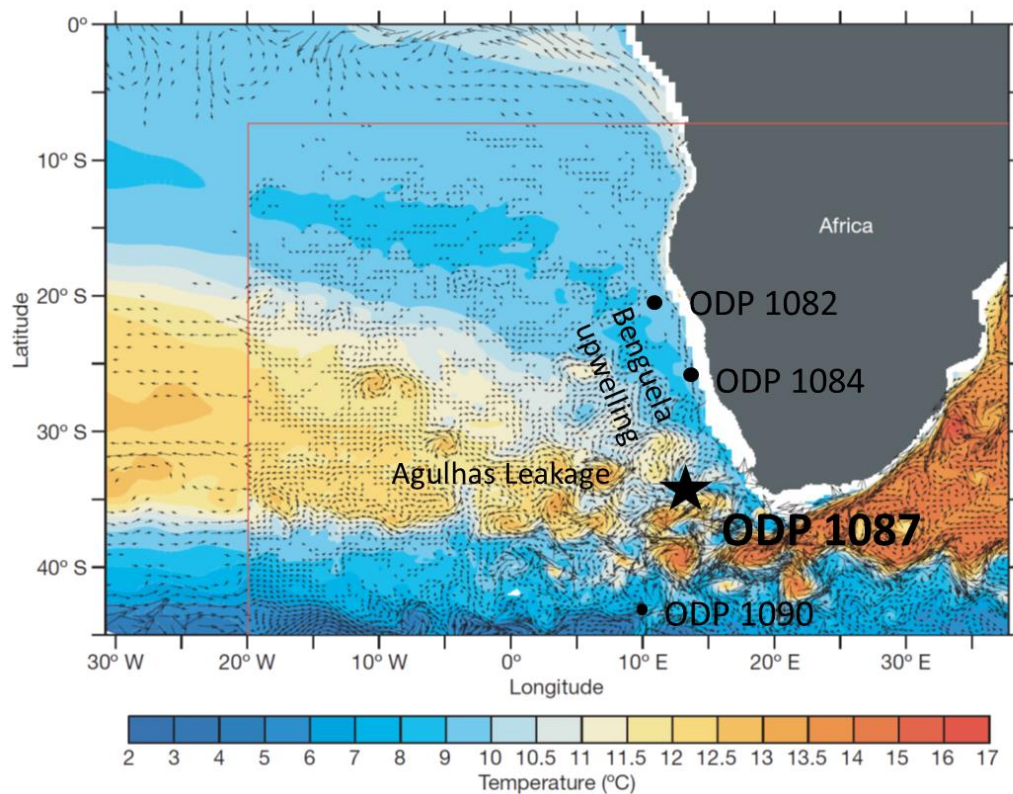


Figure 1. Location map. Location of the core site (ODP 1087) on a SST and surface ocean map of SST (Colors) and current strength and position (arrows) (Biaosoch et al., 2008). The location of local Pliocene records and the location of major oceanic systems in their modern day positions are shown.

Here, we aim to explore changes in the local hydrography in the southeast Atlantic Ocean during the Late Pliocene, by reconstructing SSTs, export productivity, and surface water masses from 3.5 to 3.0 Ma at Ocean Drilling Program (ODP) site 1087, which is located offshore of the modern southern Benguela Upwelling system (Figure 1). We apply two biomarker proxies to estimate sea surface temperatures (SSTs): the $U^{K_{37}}$ and TEX_{86} indices (Müller et al., 1998; Schouten et al., 2002). Chlorin pigment concentration mass accumulation rates (MAR) (Rosell-Melé and Maxwell, 1996) and MAR of alkenones (Marlowe et al., 1984; Volkman et al., 1980) are used to identify changes in export productivity. Planktonic foraminifera assemblages are used to track the presence of Benguela and/or Agulhas waters, as each water mass has distinctive species (Lee et al., 2008; Peeters et al., 2004; Ufkes and Kroon, 2012; Ufkes et al., 2000). Stable isotope ratios ($\delta^{18}O$, $\delta^{13}C$) of benthic foraminifera and $<225 \mu m$ fraction carbonate provide insight into the structure of the water column. In combination, these proxies allow us to investigate the signature of Benguela Upwelling and/or Agulhas Leakage to the Southeast Atlantic during the Pliocene.

2 Methods

2.1 Site description

ODP Site 1087 (31°28'S, 15°19'E, 1374 m water depth) was drilled during ODP Leg 175, the goal of which was to investigate the history of the Benguela Upwelling system. However, Site 1087 was drilled south of the major upwelling cells, with the aim of examining changes in the Agulhas Leakage (Shipboard Scientific Party and Party, 1998). In this study, samples were taken every 12 cm throughout the Late Pliocene, using the initial shipboard age model (Shipboard Scientific Party, 1998). The age model was subsequently refined using the stable isotope data, as described below, and samples were then taken every 4 cm between 3.5 and 3.0 Ma, to achieve an average sample resolution of 3 kyr.

2.2 Biomarkers

The biomarkers (alkenones, glycerol dialkyl glycerol tetraethers or GDGTs, and chlorin pigments) were extracted from homogenised, freeze-dried sediment using a CEM microwave system with 12ml of DCM:MeOH (3:1, v/v). Internal standards were added for quantification (5 α -cholestane, dotriacontane and tetracontane). The microwave temperature programme heated samples to 70°C over 5 minutes, held at 70°C for 5 minutes and then cooled down over 30 minutes (Kornilova and Rosell-Mele, 2003). The supernatant was then decanted into vials, and the extracts were dried under a gentle stream of nitrogen. An aliquot was taken for chlorins and GDGTs analyses. The remainder was derivatised using N,O-bis(trimethylsilyl)trifluoroacetamide with trimethylchlorosilane at 70°C for 1 hour prior to be analysed to quantify alkenones using a gas chromatograph fitted with a flame-ionisation detector (GC-FID) and a 30m HP1-MS capillary column. The injector temperature was held at 300°C, and the detector at 310°C. The oven program was as follows: after injection, hold at 60 °C for 1 min, increase to 120°C at 20°C m⁻¹, to 310°C at 6°C m⁻¹, and hold at 310°C for 30 min. The U^K_{37'} was calculated using the relative abundances of the C_{37:3} and C_{37:2} alkenones (Prahl and Wakeham, 1987), and converted to SSTs using the Müller et al. (1998) calibration. The total alkenone concentrations were determined by reference to the internal standard mass, and then converted to alkenone MAR using the new linear sedimentation rates and the shipboard dry bulk density measurements (Shipboard Party 1998), following the approach of Emeis et al. (1995).

Based on the results from the U^K_{37'} analysis, a subset of 20 samples was selected for further investigation using the TEX^H₈₆ index. An aliquot of the total extract was separated into different compound classes using silica column chromatography (5% H₂O) and eluted with solvents of increasing polarity: Fraction 1 (hexane), Fraction 2 (dichloromethane; alkenones) and Fraction 3 (methanol; GDGTs). All fractions were dried under a stream of N₂. Fraction 3

was re-dissolved in 200 μ L of hexane:*n*-propanol (98.5:1.5, v/v) and an internal standard (GR) was added to the sample, which was filtered through a 0.5 μ m PTFE filter. The filtered samples were analysed by high performance liquid chromatography - mass spectrometry (HPLC-MS), using a Dionex P680 HPLC coupled to a Thermo Finnigan TSQ Quantum Discovery Max quadrupole MS, with an atmospheric pressure chemical ionization (APCI) interface set in positive mode. Instrumental and chromatographic conditions were adopted from Fietz et al. (2011). GDGTs were detected in selected ion monitoring (SIM) mode at the following mass to charge ratio (*m/z*): 1302, 1300, 1298, 1296, 1292, 1050, 1048, 1046, 1036, 1034, 1032, 1022, 1020, 1018, and the internal standard GR at 1208 *m/z*. TEX^H₈₆ was calculated using the relative abundances of GDGT I, GDGT II, GDGT III and Crenarchaeol based on the methodology by Kim et al. (2010).

Chlorins were analysed using an HPLC system coupled to a photo-diode array spectrophotometer (PDA) in offline mode. An aliquot of the original solvent extract was dissolved in 2 mL of acetone, and 40 μ L was injected into the HPLC. The absorbance at the 410 nm wavelength was used to calculate, in triplicate measurements, mean chlorin pigments concentration per sample. Analytical variability was monitored using repeated measurements of a standard, and was determined at 0.07 absorbance units. Finally, for all samples, the chlorin concentration at 410 nm was divided by the original weight of the total sample to get a value of absorbance per gram; chlorin pigments MAR was calculated following the approach outlined above for alkenone MAR.

2.3 Foraminifera assemblages and stable isotopes

For foraminiferal stable isotope analysis, 80 samples (12 cm resolution) of approximately 5 cc of sediment were washed through a 150 μ m sieve. The >150 μ m fraction was dried in a 40°C oven and examined under a binocular microscope to identify and pick 3–8 *Cibicidoides wuellerstorfi* benthic foraminifera (BF) tests.

Due to a lack of *C. wuellerstorfi* tests in some samples, stable isotope analysis on <225 µm fraction carbonate (fine fraction) were undertaken, using a second set of 87 5 cc samples at 4 cm resolution. The <225 µm fraction (mostly represented by coccoliths) was collected using 0.5 µg sieves.

Approximately 30-100 µg of carbonate (fine fraction FF, or benthic foraminifera BF) were analysed for stable isotope ratios using an IsoPrime dual inlet mass spectrometer plus Multiprep device. Isotope ratios ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) are reported as per mille (‰) deviations of the isotopic ratios ($^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$) calculated to the VPDB scale using a within-run laboratory standard (KCM) calibrated against NBS-19. Analytical reproducibility of KCM is < 0.1‰ for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

The >225 µm fraction was retained and 14 samples were selected for planktonic foraminifera analysis, guided by the $\text{U}^{\text{K}_{37}}$ and chlorin data. The assemblage samples were passed through a 125 µm sieve. Because of the location of ODP site 1087 is close to the polar front and Benguela upwelling the fraction >125 microns, rather than >150 microns was used to pick foraminifera. In polar and upwelling regions, it is better to study assemblages in the >125 microns to cover all adult forms of planktonic foraminifera, as they tend to be smaller in cold waters masses (Stephan Steinke, MARUM, personal communication in 2011). The samples were also chosen to replicate studies that had been previously done at this site (Giraudeau, 1993; Giraudeau et al., 2001; Pierre et al., 2001). For each sample, 300 to 410 foraminifera were picked in order to reflect the total diversity of the foraminifera. Percentages of individual species vs. total number of foraminifera counted were calculated and subcategories of abundances formed as follows: dominant (>40% of total assemblage), abundant (20-40%), few (10-20%), rare (4-10%), present (<4%).

3. Results

3.1 Age model

According to the shipboard age models (Shipboard Scientific Party, 1998), the Late Pliocene occurred after 147 mcd, which was located using the Mammoth magnetic reversal at 3.33 Ma. The low resolution (12 cm) isotopic analysis of the benthic foraminifer *C. wuellerstorfi* ($\delta^{18}\text{O}_{\text{BF}}$) was able to identify several key excursions, as detailed in the benthic $\delta^{18}\text{O}$ stack, LR04 (Lisiecki and Raymo, 2005), including ‘KM2’, which is considered to be one of the coldest stages during the Pliocene (De Schepper et al., 2009; Dwyer and Chandler, 2009; Passchier, 2011) (Figure 2c). To supplement the $\delta^{18}\text{O}_{\text{BF}}$ record, we followed the approach of Dickson et al. (2010), developed using a proximal core (ODP Site 1085), by using $\delta^{18}\text{O}_{\text{FF}}$ (carbonate <225 μm) (Figure 2d). The $\delta^{18}\text{O}_{\text{FF}}$ record includes changes in surface ocean temperature and salinity; however, Dickson et al. (2010) showed that there is a close relationship between $\delta^{18}\text{O}_{\text{FF}}$ and $\delta^{18}\text{O}_{\text{BF}}$ in the Southern Benguela region, which allows for orbitally-tuned age model construction, especially where there is a fragmentary $\delta^{18}\text{O}_{\text{BF}}$ record.

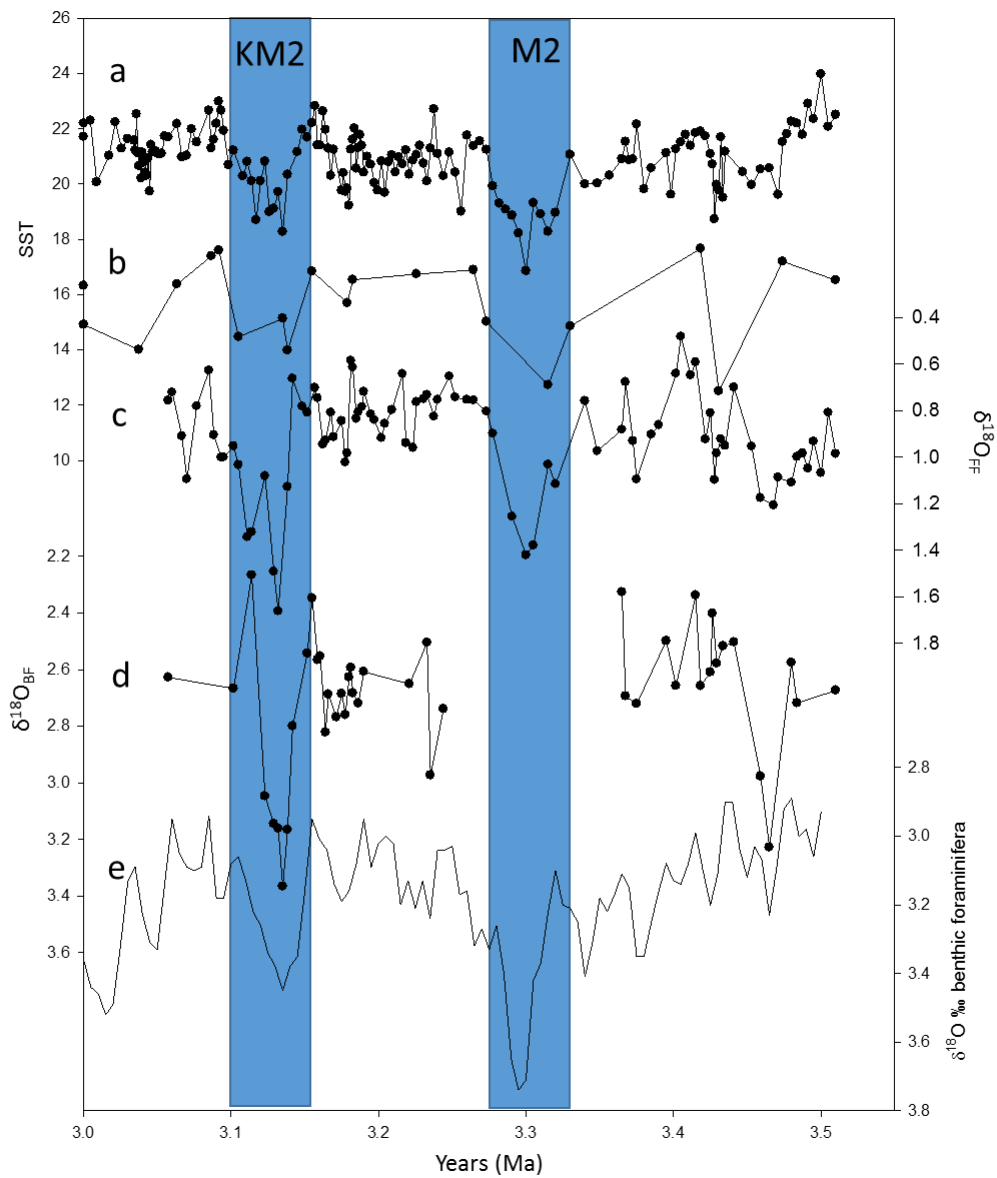


Figure 2. Late Pliocene temperature and stable isotope records from ODP Site 1087 plotted against the new benthic $\delta^{18}\text{O}$ stratigraphy. The two major cold periods in the records are marked by vertical blue bars. a) $U^{K_{37}}$ SSTs calculated using the calibration of Muller et al. (1998), b) $\text{TEX}^{\text{H}}_{86}$ temperatures determined using the calibration of Kim et al. (2010); c) $\delta^{18}\text{O}_{\text{FF}}$; d) record of $\delta^{18}\text{O}_{\text{BF}}$ re-tuned to the LR04 global benthic $\delta^{18}\text{O}$ stack (Lisiecki and Raymo, 2005); and e) LR04 global benthic $\delta^{18}\text{O}$ stack.

The $\delta^{18}\text{O}_{\text{FF}}$ record from ODP Site 1087 was visually tuned to LR04 using the two major cold periods (M2, KM2) as tie points to fix the age model (Figure 2). The age uncertainty in the tuning component of the age model is ~ 10 ka based on multiple tuning attempts for the finer scale tuning. The comparison between the $\delta^{18}\text{O}_{\text{FF}}$ and $\delta^{18}\text{O}_{\text{BF}}$ shows that the offset

between the two records (where enough data existed to compare) is at most 4 ka, although for most of the time it is less than 1 ka, and there is no consistent lead or lag between them. The timing of the large positive excursion in benthic $\delta^{18}\text{O}$ marking the “M2 glaciation” at 3.30 Ma is close to the timing of the Mammoth magnetic excursion previously identified at 147 mcd at 3.33 Ma (Shipboard Scientific Party, 1998), giving confidence in the $\delta^{18}\text{O}$ age model. The remainder of the record was tuned to LR04 using both the $\delta^{18}\text{O}_{\text{BF}}$ and $\delta^{18}\text{O}_{\text{FF}}$ data.

3.2 Temperature

The $U^{K_{37}}$ values between 3.0 and 3.5 Ma range between 0.59 and 0.83, which convert to an SST range of 17 to 22.5°C (mean SST is 21°C) (Figure 2a). There are two prominent cold periods in the record. The first occurs between 3.32 and 3.26 Ma (the M2 stage), when SSTs are between 16.5 and 19°C. The second major cooling occurs between 3.15 and 3.10 Ma (KM2), with SSTs ranging between 18 and 21°C. $\text{TEX}^{\text{H}}_{86}$ temperatures ranged from 13 to 17.5 °C with a mean of 15 °C (Figure 2b). The $\text{TEX}^{\text{H}}_{86}$ temperature minima are found at 3.4 Ma (13 °C), during M2 (13 °C) (3.3 Ma) and around 3.05 Ma (14 °C) during KM2. There is a discrepancy of approximately °C. This offset is consistent throughout the entire record.

3.3 Other biomarkers

The mean total alkenone concentration is 25.53 $\mu\text{g g}^{-1}$ (range: 14.9-51.5 $\mu\text{g g}^{-1}$). The average alkenone MAR is 2.43 (range: 0.67-5.59 $\mu\text{g (yr cm}^2)^{-1}$; Figure 3b). The three highest values occurred before 3.30 Ma, when the range of values is high. After 3.30 Ma, average alkenone MAR falls, and the range of MAR values is much reduced.

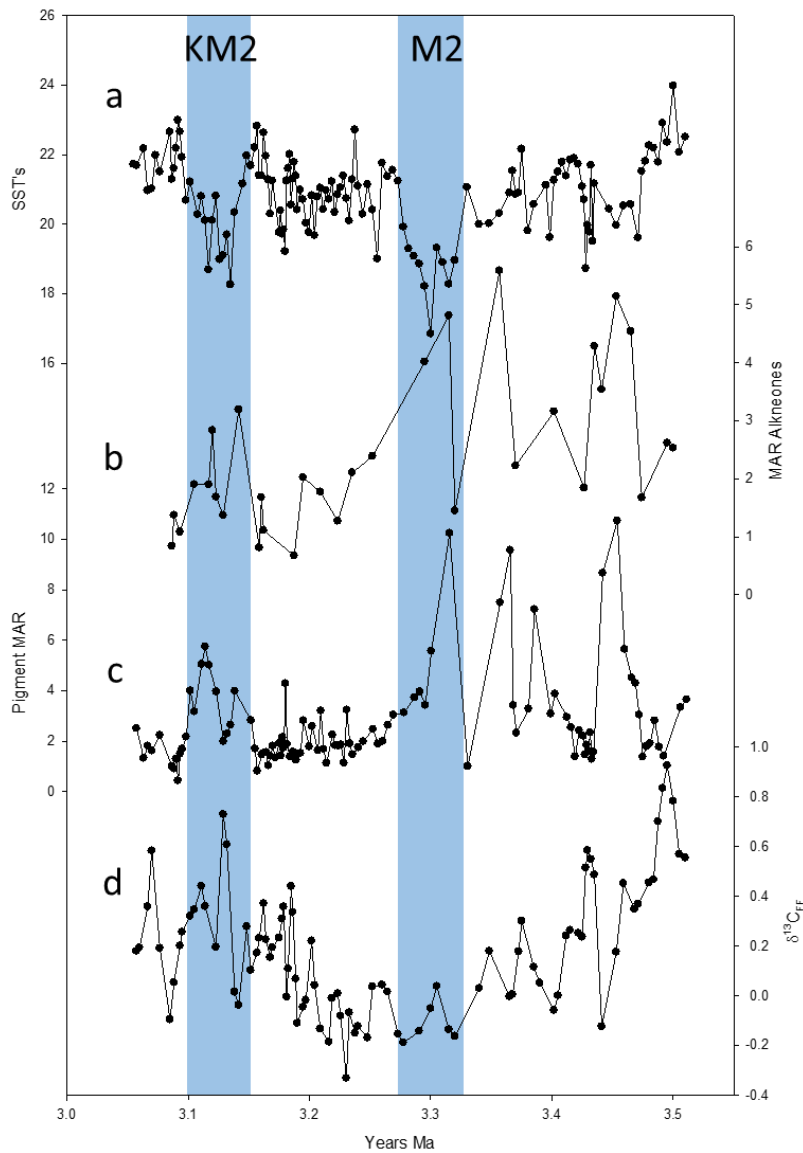


Figure 3 SSTs and indicators of export production from ODP Site 1087. The two major cold periods are marked by blue bars. a) $U^{K_{37}}$ SSTs calculated using the calibration of Muller et al. (1998); b) Total C_{37} alkenone mass accumulation rates (MAR); c) Chlorin pigments MAR (410 nm). d) $\delta^{13}C_{<225 \mu m \text{ fraction}}$.

Total chlorin concentrations range between 0.09 and 0.73 abs g^{-1} ; chlorin MAR range between 0.4-18.25 $\text{g}^2 (\text{yr cm}^2 \text{abs})^{-1}$. There is greater variability and a higher overall mean in chlorin concentration and MAR prior to 3.30 Ma. In general, chlorin and alkenone MARs increase when $U^{K_{37}}$, TEX^H_{86} indices indicate cold SSTs. (Figure 3c).

3.5 Stable isotope records: <225 μm fraction and benthic foraminifera carbonate

The mean $\delta^{18}\text{O}_{\text{FF}}$ is +1.0‰ (range = +1.6‰ to +0.5‰, Figure 2c). There are two increases in $\delta^{18}\text{O}_{\text{FF}}$ centred on 3.3 and 3.1 Ma, where $\delta^{18}\text{O}$ exceeds +1.4‰. The mean $\delta^{13}\text{C}_{\text{FF}}$ is 0.2‰ (range = +0.9‰ to -0.3‰, Figure 3d). The $\delta^{13}\text{C}_{\text{FF}}$ shows a long-term decrease by 1 ‰ from 3.5 to 3.3 Ma. The $\delta^{13}\text{C}_{\text{FF}}$ then increases by around 0.7‰ from 3.3 to 3.0 Ma. It has been shown that the fine fraction carbonate at the site is dominated by coccoliths (McClymont et al., 2005), so it is assumed that the <225 μm fraction record is dominated by coccoliths and that the isotopes reflect changes from the original coccolithophores. The influences on the $\delta^{13}\text{C}_{\text{FF}}$ are complex and there are few $\delta^{13}\text{C}$ studies on coccoliths (Ziveri et al., 2003). However, the few studies that have been done seem to indicate a higher $\delta^{13}\text{C}$ signature during times of higher productivity (Malinverno et al., 2008; Ziveri et al., 2003). This also fits the interpretation of the $\delta^{13}\text{C}$ from foraminifera previously studied at this site (Pierre et al., 2001). Therefore it is assumed that at least part of the $\delta^{13}\text{C}_{\text{FF}}$ record is influenced by changes in productivity.

The $\delta^{18}\text{O}_{\text{BF}}$ shows broadly similar trends to the $\delta^{18}\text{O}_{\text{FF}}$ (Figure 2d). There are positive isotope excursions centred on 3.1 Ma and 3.45 Ma. The gap in the $\delta^{18}\text{O}_{\text{BF}}$ around 3.3 Ma, which lasts 115 ka, is due to the lack of *C. wuellerstorfi* preserved in this part of the record.

3.6 Planktonic foraminifera assemblages

The mean percentages of the most common assemblages planktonic foraminifera change from 3.24 onwards (Table 1). Before 3.24 Ma, the majority of the identified planktonic foraminifera are *Globigerinoides bulloides*, which is always either abundant or dominant, and *Globorotalia inflata*, which is mainly classified as abundant. The only time when these two species do not dominate the planktonic foraminifera assemblage is in the sample dated 3.25 Ma; here, *G. inflata* is classified as rare although *G. bulloides* is still abundant. In contrast to the other

samples, *Neogloboquadrina pachyderma* (sin) is abundant in the sample at 3.25 Ma; although it is usually less than 10% of the assemblage (few or rare abundance) for the remaining Pliocene samples analysed here.

Age/ Species	3076.7	3085.0	3121.7	3135.0	3163.6	3170.7	3181.0	3217.0	3241.0	3283.9	3366.7	3403.2	3426.4	3467.5
<i>G. inflata</i>	A	A	A	A	F	A	D	F	A	R	D	D	A	A
<i>G. glutinata</i>	R	F	A	A	F	F	F	A	F	F	R	F	A	R
<i>G. bulloides</i>	A	A	A	A	A	A	A	A	D	A	A	D	A	D
<i>G. ruber</i>	R	P	x	R	P	x	R	P	P	P	R	R	R	F
<i>N. pachy. (dex)</i>	A	A	D	D	D	A	A	D	R	R	R	R	P	P
<i>N. pachy. (sin)</i>	F	R	R	R	F	R	R	R	F	A	R	R	F	F
<i>G. siphonifera</i>	P	P	x	P	P	x	x	P	P	P	P	x	P	x
<i>G. falconensis</i>	R	R	R	R	P	R	x	R	R	R	R	R	F	R
<i>O. universa</i>	P	R	P	x	P	R	x	P	x	x	P	P	P	P
<i>G. menardii</i>	P	R	x	R	P	x	x	x	x	P	P	x	x	P
<i>G. conglobatus</i>	x	R	P	x	P	P	R	x	R	x	x	x	P	x
<i>G. quinqueloba</i>	P	x	R	R	x	P	X	R	x	P	x	x	P	x
<i>G. sacculifer</i>	x	x	x	X	x	x	X	x	x	P	x	x	x	x

Table 1. Showing the abundance of major foraminifera through the record. Percentages of individual species vs. total number of foraminifera counted were calculated and subcategories of abundances formed: dominant (>40% of total assemblage D), abundant (20-40% A), few (10-20% F), rare (4-10% R), present (<4% P) and absent (X).

After 3.24 Ma, the majority of foraminifera are comprised of *G. bulloides*, *G. inflata*, and *N. pachyderma* (dex). The abundant or dominant classification of *N. pachyderma* (dex) after 3.24 Ma is in contrast to its much lower contribution to the assemblage prior to 3.24 Ma. Minor contributions from several other species are noted throughout the Pliocene. *Globigerinoides ruber* is composed of less than 5% of the Pliocene assemblage except in the oldest sample. A small number of warm water (Indian Ocean) species (Peeters et al., 2004) occur in the Pliocene record from ODP Site 1087: *Globigerina falconensis* is present throughout, but is usually categorised as few (<5%). There are 7 samples where *Globorotalia menardii* are identified, although there are never more than 4 individuals per sample. *G. menardii* is another important Indian Ocean warm water species, and has been identified in

ODP Site 1087 and other southeast Atlantic sediment cores during the late Pleistocene (Caley et al., 2012; Peeters et al., 2004; Sexton and Norris, 2011)

4. Discussion

4.1 Pliocene conditions at ODP Site 1087

At ODP Site 1087, the Late Pliocene is marked by an overall absence of any long-term trends in SST, $\delta^{18}\text{O}_{\text{FF}}$ and $\delta^{18}\text{O}_{\text{BF}}$. However, there are trends in the alkenone and chlorin MAR, which show shifts in variability halfway through the record at 3.30 Ma, and in $\delta^{13}\text{C}_{\text{FF}}$, which shows an initial decreasing record and then an increasing record after 3.30 Ma. On orbital and shorter timescales, the general relationships between the records shown in Figures 2 and 3 are as follows: high SSTs ($\text{U}^{\text{K}_{37'}}$ and $\text{TEX}^{\text{H}_{86}}$) are usually associated with low chlorin MAR, low alkenone MAR and lower $\delta^{18}\text{O}_{\text{FF}}$, $\delta^{18}\text{O}_{\text{BF}}$ and relatively lower $\delta^{13}\text{C}_{\text{FF}}$. Periods of lower SST ($\text{U}^{\text{K}_{37'}}$ and $\text{TEX}^{\text{H}_{86}}$) are associated with higher chlorin and alkenone MAR and higher $\delta^{18}\text{O}_{\text{FF}}$ and $\delta^{18}\text{O}_{\text{BF}}$.

The most pronounced excursion in all of the data sets occurs during the time of the M2 glacial stage c. 3.30 Ma (Figures 2 and 3). There is a 5 °C drop in $\text{U}^{\text{K}_{37'}}$ -SST, in tandem with increases in $\delta^{18}\text{O}_{\text{FF}}$, chlorin MAR and alkenone MAR (Figure 3). Lower variability in chlorin MAR and alkenone MAR and an increase in $\delta^{13}\text{C}_{\text{FF}}$ are also recorded after 3.30 Ma.

A second major cooling event occurs between 3.15 and 3.10 Ma, during the time of the KM2 glacial stage (Figures 2 and 3). Many of the patterns observed during M2 are also recorded here (low SST, higher export productivity). However, in contrast to M2, there is no significant change in the planktonic foraminifera assemblage either during or after the cooling event (Table 1). The $\delta^{18}\text{O}_{\text{FF}}$ shows a greater increase during this glacial period than during M2, suggesting more cooling of the water, possibly coupled with fresher water.

4.2 Benguela Upwelling at ODP Site 1087

ODP Site 1087 is located offshore of the southern Benguela Upwelling cells at the present day. This location, in combination with the influence of Agulhas Leakage, keeps at present ODP Site 1087 warmer and less productive than the main Benguela Upwelling cells to the north and east (Boebel, 2003). Warmer-than-present Pliocene SSTs at ODP Site 1087, as determined here, could imply an increased Agulhas Leakage of warm waters into the SE Atlantic. However, Late Pliocene SSTs at ODP Site 1087 are colder than both at ODP sites 1082 and 1084, which lie beneath the northern and central cells of the upwelling at the present day (Figures 1 and 4) (Etourneau et al., 2009; Marlow et al., 2000; Rosell-Melé et al., 2014). This is an unexpected scenario if the Agulhas current, and its leakage, were more vigorous during the Late Pliocene. The relatively cool SSTs observed at ODP Site 1087 in the Pliocene, compared to the northern Benguela Upwelling cells, indicate that there was not a contribution of warmer water to the site, as is observed during periods of increased Agulhas Leakage in the late Pleistocene (Peeters et al., 2004). Furthermore, two species of primarily Indian Ocean foraminifera occur in low abundances at ODP Site 1087 during the Late Pliocene: *G. falconensis* (rare/present), used to identify Agulhas Leakage by Peeters et al., (2004), and *G. menardii* (rare/present) (Table 1). The latter species is particularly important because it is considered to be indicative of Agulhas Leakage during the late Pleistocene for ODP Site 1087 (Peeters et al., 2004; Caley et al., 2012), whereas it is only present (but rare) during times of inferred lower export productivity in the Pliocene at ODP Site 1087. The data from ODP Site 1087 thus suggest a reduced influence of Agulhas Leakage to the southeast Atlantic in the Late Pliocene. It is possible that the Agulhas Leakage occurred, but that it is not recorded at ODP Site 1087 because it was located further to the south overall. However, the evidence at ODP Site 1087 supports the proposal that the vigour and scope of the Agulhas current has increased since the Miocene to the Pleistocene (Diekmann *et al.*, 2003).

The absent or weak influence of Agulhas leakage at ODP Site 1087 during the Pliocene means that the oceanographic regime in the southeast Atlantic was different from the modern. TEX₈₆^H temperatures are always colder than those recorded by the U₃₇^K' index (Figure 2) , which has also been observed in the modern Benguela Upwelling system (Lee et al., 2008). In the modern upwelling cells, the Archaea producing GDGTs are interpreted to be found deeper in the water column due to higher nutrient availability (Lee et al., 2008). More recent studies have shown that there is a connection between increased nutrient supply in areas such as upwelling systems and offsets of temperatures between the TEX₈₆ and U₃₇^K' because the increase in productivity may change the water depth at which GDGTs are predominately produced (Hernández-Sánchez et al., 2014; Huguet et al., 2007; McClymont et al., 2012; Shaari et al., 2013). The U₃₇^K'-TEX₈₆ offset during the Pliocene at ODP Site 1087 therefore may be interpreted to be a result of upwelling. Further support for the interpretation of higher amounts of upwelling comes from the higher export of chlorin and alkenones in the Pliocene compared to late Pleistocene values (Petrick 2014). Finally the foraminifera assemblage supports evidence for upwelling, given that the dominant foraminifera species (*G. inflata* and *N. pachyderma (dex)*), are prominent in the modern Benguela Upwelling system (Lee et al., 2008). *G. inflata* is indicative of the outer edges of the upwelling (Groeneveld and Chiessi, 2011), while *N. pachyderma (dex)* is found in the filaments produced by seasonal upwelling around the permanent upwelling cells (Ufkes et al., 2000; Ufkes and Kroon, 2012). The combined temperature, chlorin, alkenone, $\delta^{13}\text{C}_{\text{FF}}$ and foraminifera assemblage datasets indicate that during the Pliocene the southern Benguela Upwelling system was more extensive and/or less seasonal in nature, compared to its modern and late Pleistocene operation. Supporting our interpretation that ODP Site 1087 was affected by a southward displaced Benguela Upwelling system relative to present is that there is no expression of the M2 or of the KM2 glacial stages at the northern Benguela Site, ODP 1082 (Figure 4). This suggests that these cool, high

productivity (upwelling) events were isolated to the southern Benguela region, which was likely controlled by (or sensitive to) different processes than the northern Benguela.

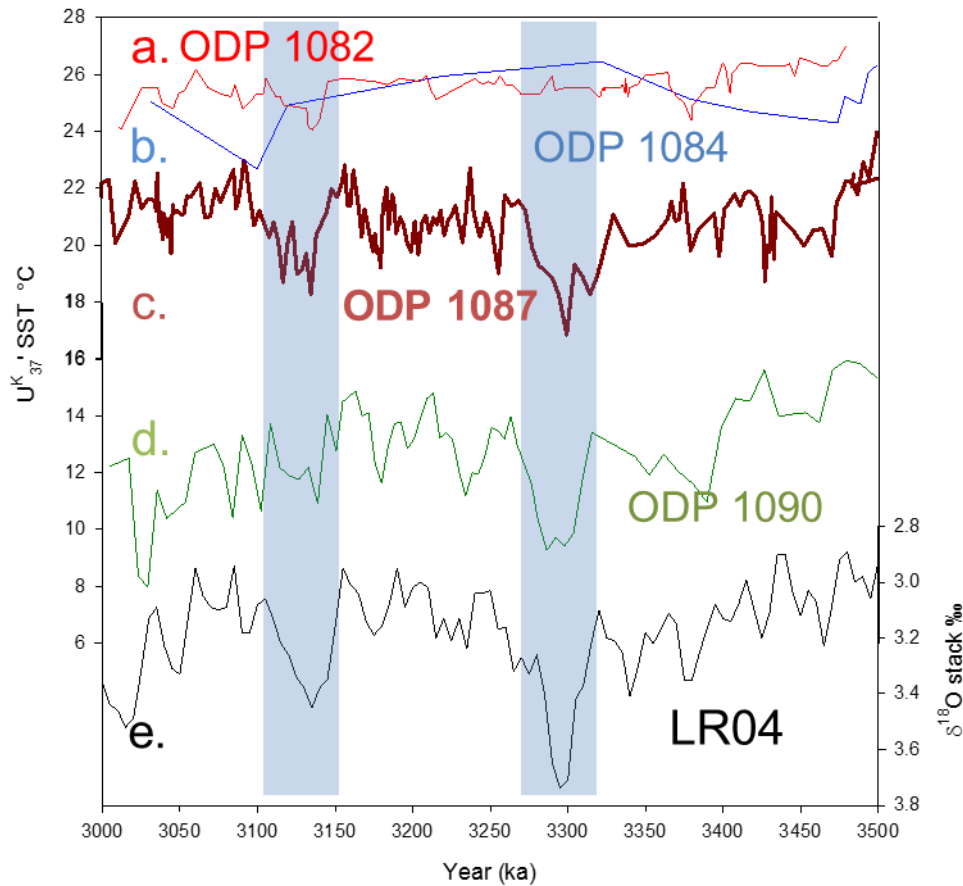


Figure 4. Late Pliocene SSTs in the southeast Atlantic. a) ODP1082 (red; Etourneau et al., 2009) b. ODP 1084 (blue), (Marlow et al., 2000) c) ODP 1087 (dark red) (This study) d) ODP 1090 (Dark green) (Martinez-Garcia et al., 2011). e) The LR04 Benthic stack (Black) is shown for reference (Lisiecki and Raymo, 2005). Blue bars represent cold periods.

4.3 Changes in the Southeast Atlantic during the Pliocene

Decreasing SSTs and increasing productivity have been used to argue that the northern and central upwelling cells of the Benguela system developed their modern form around 2.4-2.0 Ma (Etourneau *et al.*, 2009,2010, 2012). The development of an SST gradient between

ODP sites 1081 and 1084 from 3.5 Ma (Figure 4) was interpreted by Rosell-Mele et al. (2014) to indicate an earlier initial onset of Benguela Upwelling intensification (although still weaker than modern), and was closely linked to increased dust accumulation in the Subantarctic Atlantic, which reflects intensification of atmospheric circulation at this time (Martinez-Garcia et al., 2011). Furthermore, recent studies show that upwelling may have been present in the northern upwelling sites between 3.5 and 3.0 Ma; however, it was suggested that there was a fundamental difference between Pliocene upwelling and modern day upwelling based on the change in the biomarkers seen in the early Pliocene (Leduc et al. 2014). Rosell-Mele et al. (2014) interpreted the low SST gradient between ODP 1081 and 1084 to indicate a warm, persistent ‘Benguela El Niño-like’ state for the region during the Pliocene (Rosell-Melé et al. 2014). Our results support this interpretation in principle, demonstrating that active upwelling was present in the Pliocene, but that it was displaced southwards into the Southern Benguela region relative to modern. This interpretation contrasts with earlier studies that have assumed there was little Benguela Upwelling occurring before 3.0 Ma (Brierley and Fedorov, 2010; Fedorov et al., 2007, 2010, 2006).

As a wind-driven system (Hutchings et al., 2009), the Pliocene displacement in the Benguela Upwelling is likely related to a change in the position and/or intensity of atmospheric circulation in the Southern Hemisphere compared to modern. Today, upwelling only affects the southern part of the Benguela upwelling system during the austral summer due to the southward expansion in the wind patterns, which are only seasonally active in this part of the system, whereas perennial upwelling dominates the central and northern regions (Andrews and Hutchings, 1980). However, during the Late Pliocene, it has been argued that the Hadley cells were expanded 3-4 degrees of latitude southward (Brierley et al., 2009; Etourneau et al., 2010). From 3.5 Ma, increased dust deposition to ODP Site 1090 (Martinez-Garcia et al., 2011) is interpreted to reflect an intensification of atmospheric circulation in the southern hemisphere

and/or reduced precipitation in South America and Africa, both consistent with relatively weak Hadley circulation (Brierley et al., 2009). A southward displacement of the trade winds would have been conducive to the development of perennial upwelling in the southern Benguela region (Christensen et al., 2002). The dominant proposal to explain the shifting and/or strengthening Hadley circulation over the Pliocene and Pleistocene draws on the strengthening of meridional temperature gradients which have evolved over this time window, particularly clearly expressed in SST data sets (Brierley et al., 2009; Fedorov et al., 2010, 2006; Martinez-Garcia et al., 2010). However, model sensitivity tests have recently demonstrated that the intensity of the Benguela upwelling system could be impacted by tectonic uplift in the African continent through the Miocene and Pliocene (Jung et al., 2014). For example, an intensification of Ekman pumping in the Benguela region of up to 40-60% was demonstrated, due to altered atmospheric circulation in response to uplift in southern and eastern Africa, but the potential for strong regional differences was also indicated, e.g. between northern and southern Benguela regions (Jung et al., 2014). It is currently difficult to assess whether this mechanism can explain a southward displacement of the Benguela upwelling during the late Pliocene, as we suggest here, since the timing and pattern of uplift in Africa remains controversial (see discussion by Jung et al., 2014) and there is evidence for gradual cooling (and inferred upwelling intensification) in the Peru upwelling system since 4 Ma (Dekens et al., 2007), which could not be explained through an African uplift mechanism. Further work is required to constrain the meridional SST gradients outside of the upwelling regions, as well as other indicators of Hadley circulation changes, alongside improved chronologies of African uplift, in order to better assess the relative importance of these different mechanisms in explaining the patterns observed in the Pliocene Benguela upwelling system.

4.4 Climate changes in the Southern Hemisphere from 3.30 Ma

The decreases in chlorin MAR and alkenone MAR after the M2 glacial stage suggest a change in the nature of the Benguela upwelling system through the late Pliocene (Figures 3). M2 also marks an inflection point in the $\delta^{13}\text{C}_{\text{FF}}$ record, which shows a long-term rise after M2. The planktonic foraminifera species distribution shows major changes during and after the M2 glacial stage. This is especially true of *N. pachyderma* (*sin.*), which shows a peak during M2 and *N. pachyderma* (*dex.*), which increases after 3.3 Ma (Table 1). Both of these species are associated with the Benguela upwelling in the modern system (Giraudeau et al., 2002, 2001; Ufkes and Kroon, 2012; Ufkes et al., 2000). While this is based on low resolution samples, it fits the patterns seen in the higher resolution geobiological data, by showing a major shift in the environment of the upwelling after M2. The M2 stage at ODP Site 1087 is therefore important in marking a shift in the nature of the organic deposition at ODP Site 1087.

The M2 glacial stage has also been described as being an important marker of long-term shifts in southern hemisphere climate (Lisiecki and Raymo, 2007; McKay et al., 2012). The ANDRILL core, recovered in the Ross Sea sector of Antarctica, suggests that M2 is the start of major cooling in Antarctica, and that this cooling is coupled with a greater instability in the ice sheets (McKay et al. 2012). Antarctic ice-sheet growth around M2 is linked to the beginning of expansion of the ice shelves around Antarctica (McKay et al., 2012; Riesselman and Dunbar, 2013). Furthermore, a major shift in the diatom assemblage occurs just after M2, with the first appearance of a number of cold water species (Riesselman and Dunbar, 2013). These patterns suggest that conditions in Antarctic and in the Southern Ocean were shifting towards a colder, more glaciated state through the late Pliocene, concomitant with the shift in export productivity that we observed at ODP Site 1087.

However, we do not observe any cooling trend at ODP Site 1087 between 3.5 and 3.0 Ma (Figure 2), nor in sites from the northern Benguela (Rosell-Melé et al. 2014) or Subantarctic Atlantic (Martinez-Garcia et al., 2010). This is opposed to the conclusions outlined in Karas et

al (2011). In that article, it was suggested that cooling in the southeast Atlantic Ocean was caused by cooler waters being sent through the Indonesian Throughflow, which were then translated through the Agulhas leakage, as was proposed by (Cane and Molnar, 2001). However, our work shows that there was no evidence of leakage influence at this time and no evidence of SST cooling between 3.5 and 3.0 Ma. This does not mean that changes in the Indonesian Throughflow did not play a role, but they do not seem to have affected the record between 3.5 and 3.0 Ma. It is still possible that the lack of an Agulhas Leakage at the site might be linked to changes in the Indonesian Throughflow, but this would need more sites to test.

As a result, the shifts in export productivity at ODP Site 1087 seem unlikely to have been controlled by physical changes in surface ocean hydrography. Rather, we hypothesise here that the changes in export productivity could have been driven by changes to the nutrient content of the waters upwelling at ODP Site 1087 through the Pliocene. This change could have been either related to the total amount of nutrients delivered to the sea surface, or to the actual components of the nutrients delivered to the site, as has been suggested for later periods in the Benguela Upwelling where an increasing amount of Fe being delivered the subsurface waters to the system in response to changing ice cover at the poles. The change in Fe led to major changes in the productivity of the ocean (März et al., 2013). This resulted in an increase in diatom production worldwide, including major diatom blooms in the Benguela upwelling (März et al., 2013). Evidence of these diatom blooms is seen further north of the site even during the Pliocene (Leduc et al., 2014). Therefore, there is previous evidence that major changes in the system can be caused by changes in the nutrient content of the subsurface.

Since upper Antarctic Intermediate Water is a major contributor to the waters being upwelled in the Benguela system, and since it forms in close association with the position of the Antarctic Circumpolar Current, shifting sea surface conditions and plankton productivity in the Southern Ocean might plausibly have led to a change in nutrient delivery to ODP Site

1087 after the M2 glacial. However, testing this hypothesis requires further work in detailing the nutrient delivery and utilisation to Benguela sites through the Pliocene, as well as improved spatial and temporal resolution data sets of changing hydrography and productivity in the Southern Ocean, in order to explain the apparent decoupling between comparatively stable upwelling conditions (as indicated by the SST data sets) and fluctuating biological production.

5 Conclusions

Our analysis of data from ODP site 1087 in the Southern Benguela region has identified a displaced and/or strengthened upwelling system during the warmth of the Pliocene relative to the modern. Using records of SST (U_{37}^K , TEX_{86}^H), and export productivity (stable isotopes, foraminifera, alkenones and chlorins), we have shown that Pliocene SSTs remained stable and were higher than modern, but that there is a stronger signal of upwelling at this site than is observed during the late Pleistocene. We infer that the Benguela upwelling system was displaced southward relative to present, leading to more seasonal upwelling in the northern Benguela region. The Agulhas Leakage is not present in ODP Site 1087 during the Pliocene. The start of Antarctic cooling and expansion of sea ice may of caused changes to the nutrient delivery to the Site but did not immediately lead to any large-scale shifts in the local ocean circulation. The southward displacement of the main Benguela upwelling cells during the late Pliocene is consistent with the occurrence of a weaker atmospheric circulation and expanded the Hadley cells in comparison to present.

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